JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 114, D00H09, doi:10.1029/2009JD012075, 2009

Clouds and water vapor in the Northern Hemisphere summertime stratosphere

A. E. Dessler[1]

Received 16 March 2009; revised 18 August 2009; accepted 25 August 2009; published 26 November 2009.

[1] Cloud top observations from the Cloud-Aerosol Lidar Infrared Pathfinder Satellite Observations (CALIPSO) instrument and water vapor measured by the Microwave Limb Sounder (MLS) are used to study the occurrence of clouds in the Northern Hemisphere (NH) summertime lower stratosphere (20°–70°N) and their relation to water vapor. At low latitudes, clouds in the stratosphere tend to occur in regions of intense convection, while at high latitudes, there is little longitudinal preference for the clouds. In general, the 0.1% cloud top occurrence contour tends to be found ~3 km or 40–50 K of potential temperature above the tropopause. At midlatitudes, the occurrence of clouds above the tropopause is associated with enhanced water vapor, suggesting that clouds are associated with moistening events in the lower stratosphere. In the subtropics, the occurrence of clouds is associated with reduced water vapor, suggesting that clouds are associated with dehydration events. Our results are consistent with hydration or dehydration being determined by the local relative humidity. Low relative humidity allows significant evaporation of lofted cloud ice, which is thought to be the key to moistening events. High relative humidity inhibits evaporation of lofted cloud ice and encourages in situ formation of clouds that are thought to play a role in dehydration.

Citation: Dessler, A. E. (2009), Clouds and water vapor in the Northern Hemisphere summertime stratosphere, *J. Geophys. Res.*, *114*, D00H09, doi:10.1029/2009JD012075.

1. Introduction

[2] While relatively rare, clouds in the stratosphere are associated with important physical processes. In the tropics, in situ formation of clouds around the tropopause can be important in dehydration of air entering the stratosphere [e.g., *Jensen and Pfister*, 2004]. Tropical convection can also reach the stratosphere [e.g., *Alcala and Dessler*, 2002; *Gettelman et al.*, 2002; *Zipser et al.*, 2006; *Dessler et al.*, 2006b], and the convective injection of mass there, including clouds, has important implications for the chemical and physical properties of the tropical tropopause layer and stratosphere. For example, convective injection of cloud mass into the TTL may play a key role in setting the isotopic composition of water entering the stratosphere [e.g., *Dessler and Sherwood*, 2003; *Dessler et al.*, 2007]. See *Fueglistaler et al.* [2009] for a review of TTL science.

[3] It is now clear that convection can also penetrate into the extratropical stratosphere. The impacts of convection are mainly seen in the so-called lowermost stratosphere [*Poulida et al.*, 1996; *Fischer et al.*, 2003; *Hegglin et al.*, 2004; *Hess*, 2005], that part of the stratosphere with potential temperature $\theta < 380$ K [*Hoskins*, 1991]. But there is also evidence that convection can reach higher, into the overworld ($\theta > 380$ K), and have important impacts there [*Fromm and Servranckx*, 2003; *Wang*, 2003; *Livesey et al.*, 2004; *Jost et al.*, 2004], particularly in the amount of water vapor [e.g., *Dessler and Sherwood*, 2004; *Hanisco et al.*, 2007].

[4] The occurrence of clouds in the extratropical stratosphere is not as well studied as it is in the tropics. In this paper, I use a new data set of cloud top heights to quantify the occurrence of clouds here and evaluate the implications for the water vapor budget there.

2. Distribution of Cloud Tops in the Stratosphere

[5] Measurements of clouds are obtained from the Cloud-Aerosol Lidar Infrared Pathfinder Satellite Observations (CALIPSO) [*Winker et al.*, 2009] level-2 cloud layer product (versions 2.01 and 2.02). This product reports cloud layer information with a horizontal resolution of 5 km and a vertical resolution of 60 m. Nighttime measurements, obtained around 1:30 A.M. local time, are used here because of the superior ability of the instrument to detect thin clouds at night. Clearly, the lack of any diurnal cycle information is a limitation of this analysis. The focus of this work is on the Northern Hemisphere during June, July, August, and September, corresponding to the time of year when stratosphere-penetrating extratropical convection is most likely to occur.

[6] Figure 1a shows the fraction of CALIPSO observations containing a cloud top above the local tropopause, using data obtained in 2007 and 2008. Local tropopause height comes from the GEOS-5 reanalysis [*Suarez et al.*,

[1]Department of Atmospheric Sciences, Texas A&M University, College Station, Texas, USA.

0148-0227/09/2009JD012075$09.00

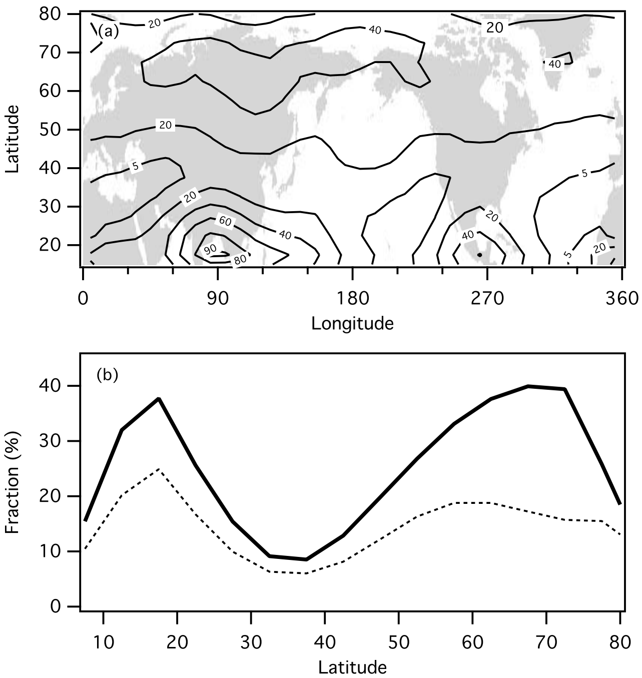

Figure 1. (a) Fraction (in percent) of CALIPSO nighttime observations that show a cloud top above the local tropopause as a function of latitude and longitude and (b) a zonal average as a function of latitude (solid line). Also shown in Figure 1b is the zonal average fraction derived from daytime measurements only (dashed line). Calculated from data obtained during June, July, August, and September 2007–2008.

2008] interpolated in time and space to the location of the CALIPSO measurement, and it is included in the CALIPSO level-2 data. The GEOS-5 tropopause is a combination of a lapse-rate tropopause at low latitudes and a dynamical tropopause at mid and high latitudes. Figure 1a shows clear maxima above the Asian monsoon and over Central America, both regions of intense convection. As latitude increases, the zonal variability decreases; at high latitudes, there is minimal variation with longitude.

[7] One quirk of the CALIPSO data is that clouds with tops *and* bottoms above the tropopause are not archived in the cloud layer product, but are instead included in the aerosol product, and thus our analysis does not include them. We have analyzed the aerosol layer product and find that the number of such clouds is much smaller than the number with tops above the tropopause and bottoms below. Most of these clouds tend to occur at low latitudes, where they would increase the cloud fraction in Figure 1 by a few percent. At mid and high latitudes, such clouds are rare and neglect of them has no impact on our analysis.

[8] One of the best data sets to compare our results to is the cloud observations made by the Stratospheric Aerosol and Gas Experiment II (SAGE II), which is described by *Wang et al.* [1996]. Figure 1a here can be compared to Plate 1 of Wang et al., and the comparison reveals good agreement. A maximum in cloud occurrence in the SAGE II data at the altitude of the tropical tropopause (16.5 km) can be found over the Asian monsoon (with SAGE II frequencies > 60%), with a lesser maximum seen by the SAGE II over Central America (~30%). At higher latitudes, the SAGE II also sees frequent clouds (~30%) in the stratosphere, and distributed somewhat evenly in longitude. Overall, the agreement between the data sets is quite good, particularly considering the differences in viewing geometry and sampling time.

[9] Figure 1b shows a zonal average of Figure 1a, showing a strong minimum around 35°N, with similar maximum values found at 20°N and 70°N. The increase in frequency from 35°N to 70°N is due primarily to the decrease in the height of the tropopause, which can be see in Figure 2, which shows the average geometric height of the tropopause, as well as heights of the surfaces of constant θ, as a function of latitude. The tropopause is found above 16 km at low latitudes, and around 10 km at 70°N. Most of the change in altitude of the tropopause occurs in a relatively narrow latitude band from 35°N to 50°N.

[10] Also shown on Figure 1b is the zonal average fraction derived from daytime measurements. This shows that nighttime data contain 50–100% more clouds than the daytime data. The shape of the latitude-longitude distribution of clouds in the daytime data (not shown) is very similar to Figure 1a. Strong day-night differences are also seen in the occurrence of tropopause-level clouds in other spaceborne lidar data sets [*Dessler et al.*, 2006a, 2006b]. Some of this difference may be real, but without more data it is impossible to separate real diurnal changes from those due to differing sensitivity of day and night measurements. It is for this reason that I've chosen to focus here on nighttime data.

[11] Figure 3 shows longitude-height slices of cloud top occurrence at a range of latitudes. Figure 4 shows the same data, but using a vertical coordinate of θ. In the lower stratosphere, θ is conserved on time scales of weeks to months [*Newman and Rosenfield*, 1997], so it is a more useful vertical coordinate for many analyses than geometric height. The θ of the cloud top is calculated using temperature and pressure profiles from the GEOS reanalysis, which are provided in the CALIPSO level-2 cloud profile product.

[12] At low latitudes, there is strong zonal variability, with the 0.1% contour extending as high as 19 km or 420 K in regions of strong convection, about 3 km or 40 K above the tropopause. In nonconvective regions, the 0.1% contour

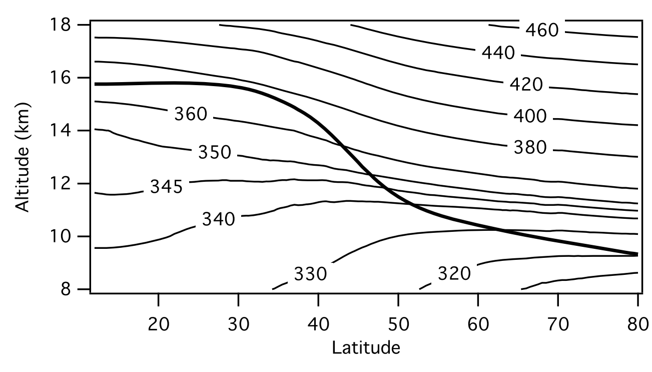

Figure 2. Tropopause height (thick line) and lines of constant potential temperature θ (thin lines) as functions of latitude. Based on GEOS reanalysis data from June, July, August, and September 2007–2008.

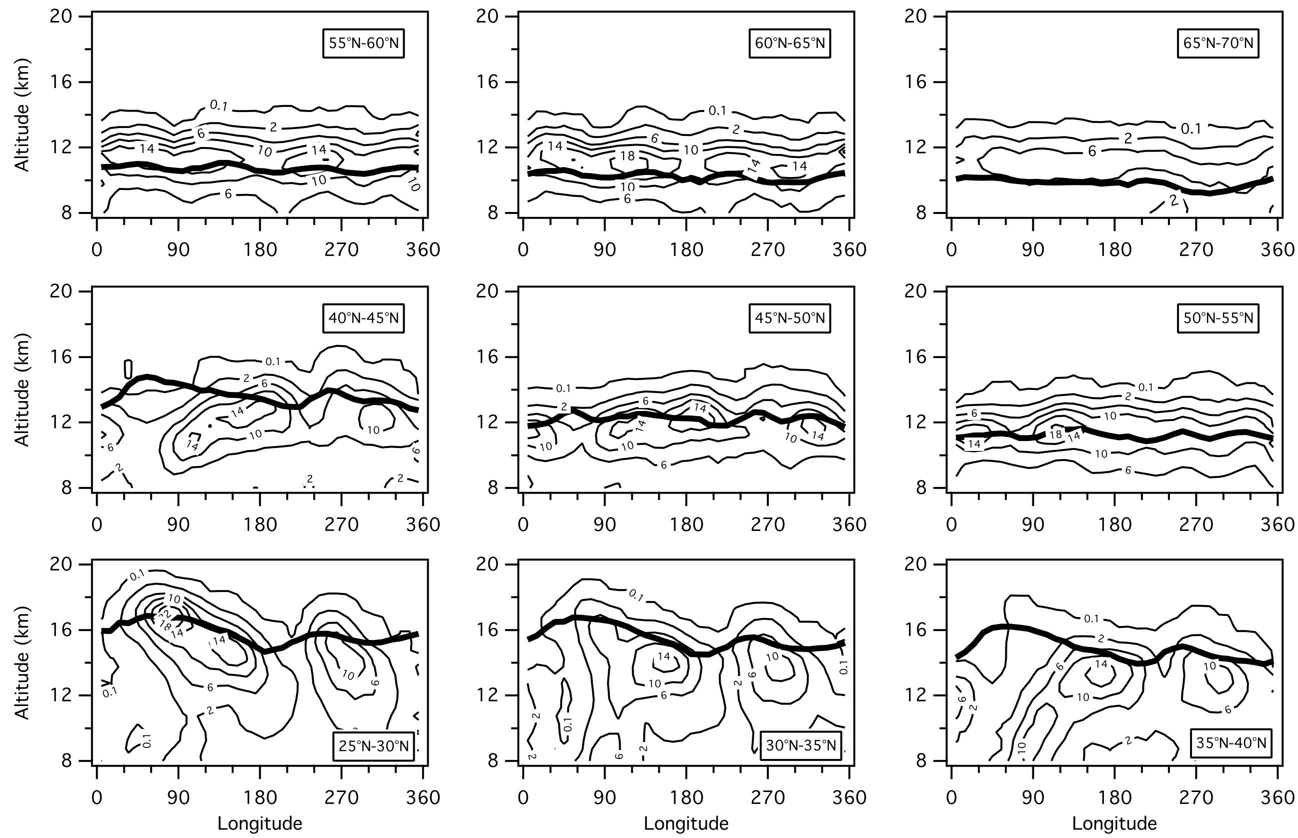

Figure 3. Percent of observations that show a cloud top, in percent/km. Each plot contains data from a single latitude range. The thick line is the average tropopause. Data were obtained from June, July, August, and September 2007–2008.

is close to the tropopause. In addition, cloud tops are higher above the Asian monsoon: the 0.1% contour is about 2 km or 30 K higher there than over Central America.

[13] Deep convection falls off rapidly with latitude in the Asian monsoon region, while dropping off more slowly with latitude over Central and North America. Thus, from 35°N to 40°N, the 0.1% contour extends to 18 km or 390 K over both Asia and North America. And from 40°N to 45°N, the 0.1% contour extends about 1 km or 15 K higher over North America than over Asia. *Dessler and Sherwood* [2004] argued that the extension of stratosphere-penetrating convection to higher latitudes over North America had important implications for the water vapor distribution in the summertime extratropical lower stratosphere, a point that will be picked up later in this paper.

[14] At higher latitudes, the zonal variability disappears. From 45°N to 70°N, the 0.1% contour is located around 14 km or 370 K at all longitudes, the 2% contour is located around 13 km or 355 K, and other contours also remain approximately fixed in altitude with longitude. This explains why the fraction of cloud tops found above the tropopause (Figure 1b) tends to increase with latitude: the distribution of cloud tops in θ remains reasonably fixed, but the tropopause descends with increasing latitude, leading to larger numbers of clouds being seen above the tropopause with increasing latitude.

[15] Figure 3 is similar to the SAGE II cloud distribution plotted in Figure 4 of *Wang et al.* [1996]. At low latitudes (20°N plot of Figure 4 of Wang et al.), there is a dominant maximum over Asia, with a lesser maximum over North America. At higher latitudes (50°N plot of Figure 4 of Wang et al.), the SAGE II data also show little longitudinal preference for cloud occurrence.

[16] Figure 5 shows latitude-height slices of cloud top occurrence for longitudes around the Asian Monsoon (Figure 5a) and around North America (Figure 5b), constructed from 2008 data (we limit the data to 2008 here to facilitate comparison with water vapor data in the next section). The two sectors show general similarities, with cloud top occurrence contours by and large following the tropopause. Figures 5a and 5b also show maxima near the low-latitude tropopause. Over Asia, this maximum is associated with the Asian Monsoon, while over North America it occurs near regions of intense convection over Central America. Figures 5a and 5b show strong similarities to the SAGE II observations (Figures 2 and 3 of *Wang et al.* [1996]), suggesting that the general features of the distribution are robust features of the atmosphere.

[17] Figure 5c shows the absolute difference between the frequencies in Figures 5a and 5b (absolute difference means that a difference between 7% and 10% is 3%), with positive anomalies corresponding to higher frequencies in the Asian sector. At low latitudes, a positive maximum centered on 17 km shows that cloud tops at this altitude are more frequent over Asia, while large negative values centered at 14 km show that cloud tops at this altitude are more frequent over Central America. This arrangement is consistent with more vigorous and deeper convection over the Asian

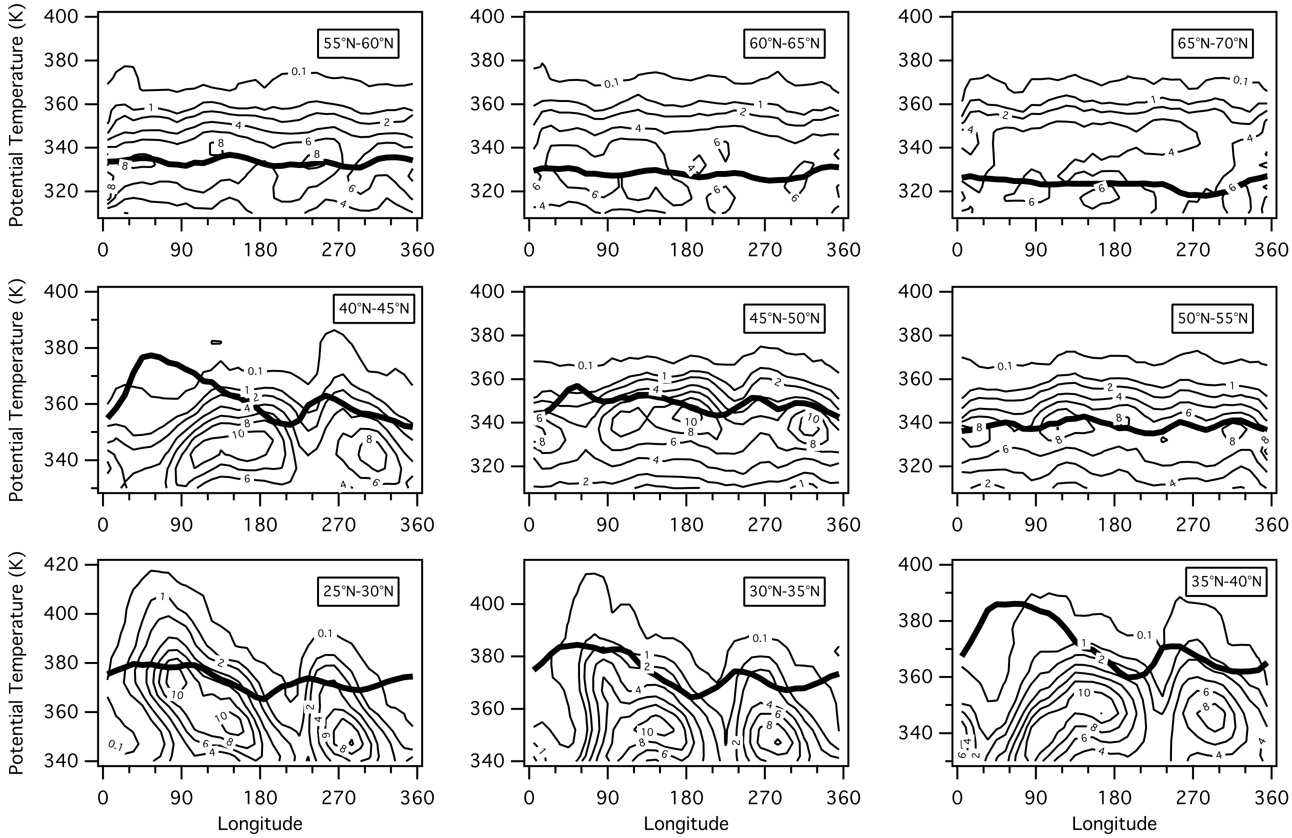

Figure 4. Percent of observations that show a cloud top, in percent/(5 K). Each plot contains data from a single latitude range. The thick line is the average tropopause. Data were obtained from June, July, August, and September 2007–2008.

monsoon [e.g., *Dunkerton*, 1995]. Given the high relative humidity at low latitudes (discussed in the next section), some of these clouds are almost certainly formed in situ, while others may be convective in origin. Determining the formation mechanism is a nontrivial exercise, and in this paper I have not made any attempt to determine the relative fractions of in situ versus convective origins.

[18] At midlatitudes, Figure 5c shows negative values in the lower stratosphere, meaning cloud top frequency is higher over North America than over Asia, as was seen in Figure 3. At high latitudes, positive values in the lower stratosphere indicate higher cloud top frequency in the Asian sector.

3. Connection to Water Vapor

[19] To look for connections between clouds and water vapor, I now analyze measurements of water vapor made during June, July, August, and September of 2008 by the Microwave Limb Sounder (MLS) on NASA's Aura satellite. During this time period, the MLS and CALIPSO measurements were made within a few kilometers and a few minutes of each other, so can be considered for all practical purposes simultaneous. Version 2.2 of the data [*Livesey et al.*, 2007] is used. Water vapor in this data set is retrieved in the upper troposphere and lower stratosphere every ~1.5 km in the vertical; the measurements have an accuracy of 10–20%, a vertical resolution of 2–3 km, and an along-track horizontal resolution of a few hundred km.

[20] Figure 6a shows zonal average water vapor mixing ratio. I have used both day and night water vapor measurements in this plot; the lifetime of water vapor at these altitudes is long enough that there is no diurnal cycle, so segregating the data into day or night data yields the same results. Like cloud top frequency, the water vapor contours tend to follow the tropopause. In general, mixing ratios decline with altitude, and the mixing ratio at the tropopause increases with latitude, mainly because the tropopause descends with height.

[21] Figure 6b shows the zonal average relative humidity (RH). RH is calculated for each MLS measurement of water vapor mixing ratio using the GEOS temperature fields interpolated to the location of the MLS measurements. Figure 6b is calculated by zonally averaging the individual RH values. Figure 6b shows that RH tends to be high in the troposphere and low in the stratosphere. Within the stratosphere, RH tends to be higher at low latitudes due to the low temperatures found there. Temperature increases with latitude in the stratosphere, and RH correspondingly decreases.

[22] Figure 6c shows the percent difference between water vapor mixing ratio over the Asian and North American sectors, with positive values indicating higher mixing ratios over Asia. Over most of the summertime Northern Hemisphere troposphere and lower stratosphere, water vapor over Asia is higher than over North America. The exception is

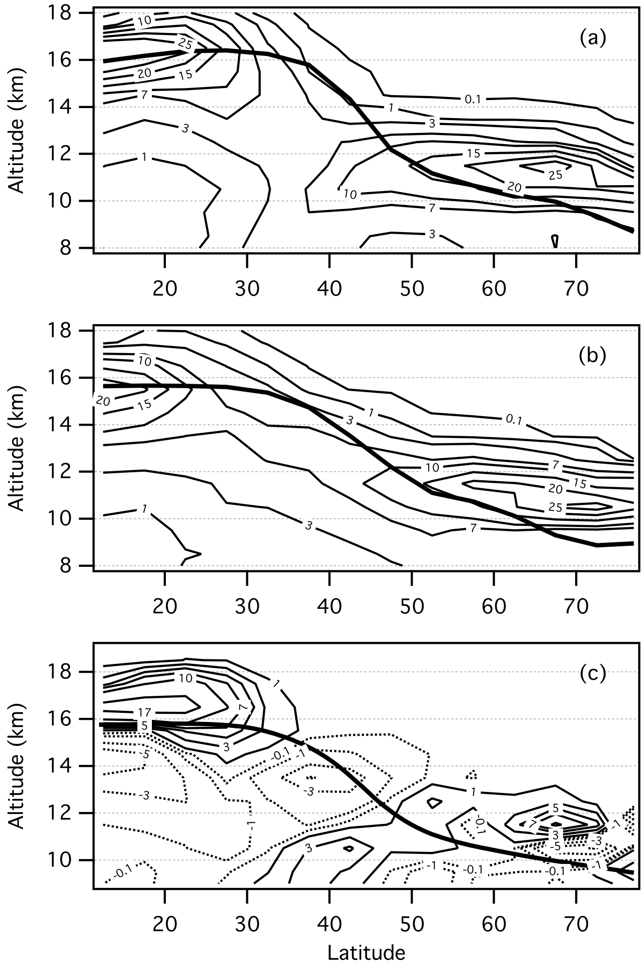

Figure 5. Cloud top fraction (in percent) and average tropopause (thick line) height as a function of latitude for (a) the Asian sector (45°–145° longitude) and (b) the North American sector (235°–305°), and (c) absolute difference between Figures 5a and 5b, with positive values corresponding to higher values in the Asian sector, and zonal average tropopause (thick line). Data were obtained from June, July, August, and September 2008.

around the midlatitude tropopause, where water vapor mixing ratios over North America are more than 20% higher than over Asia.

[23] This region of enhanced water vapor over North America matches closely the region where cloud top frequency over North America exceeds that over Asia (Figure 5c). This is consistent with the work of *Dessler and Sherwood* [2004], who concluded that enhanced convection over North America leads to injection of cloud ice into a region of low RH in the stratosphere, where cloud ice evaporates and leads to enhanced water vapor. A similar correlation between enhanced cloud occurrence and water is seen between 60°N and 70°N and just above the tropopause.

[24] In the low-latitude stratosphere, on the other hand, there appears to be little correlation between cloud top frequency anomaly (Figure 5c) and water vapor mixing ratio anomaly (Figure 6c). In fact, there are hints of an anticorrelation: collocated with the low-latitude cloud top frequency anomaly maximum is a slight negative water anomaly. One explanation for this slight anticorrelation is that the cloud top anomaly maximum is also region of high RH (Figure 6b), so while much ice is lofted by convection into this region, the high RH will prevent evaporating, and convection will therefore have little effect on water vapor mixing ratio there.

[25] In order to better understand the cause-and-effect relationship between clouds and water vapor, I have merged the MLS and CALIPSO data in order to calculate the difference between water vapor when clouds are present and absent. Because of the differing viewing geometries of the instruments, merging the CALIPSO and MLS measurements requires some care. In this analysis, for every nighttime MLS measurement, we find all CALIPSO 5 km retrievals that are within 15 km of the MLS tangent point (about six CALIPSO points) and take these to be a set of coincident measurements. Then for each coincident CALIPSO footprint, we integrate the cloud extinction over a 1.5 km

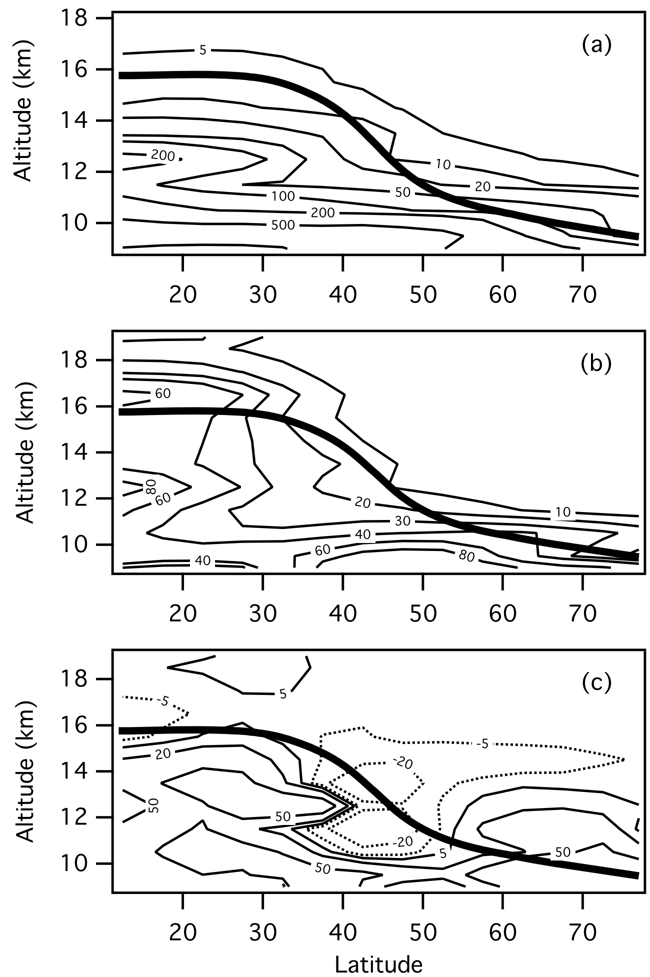

Figure 6. (a) Zonal average water vapor mixing ratio (in parts per million by volume), (b) zonal average relative humidity (in percent), and (c) percent difference in water vapor mixing ratio between the Asian and North American sectors, with positive values corresponding to higher values in the Asian sector. The thick line is the zonal average tropopause. Data were obtained from June, July, August, and September 2008.

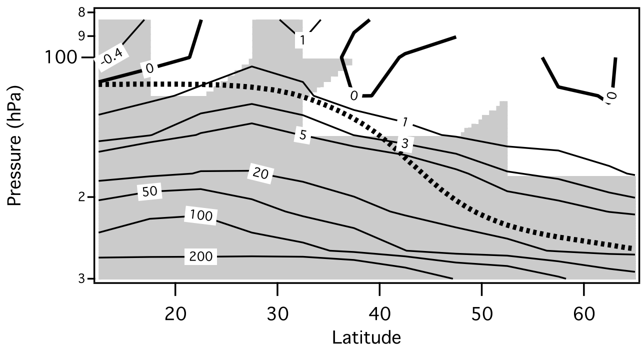

Figure 7. Contours are water vapor in the presence of clouds minus water vapor in clear skies. The gray shading indicates that the difference is nonzero at the 95% confidence level. The dotted line is the zonal average tropopause pressure.

layer centered on the tangent point pressures of the MLS measurements.

[26] A MLS measurement is considered clear sky if all of the collocated CALIPSO measurements show zero extinction. A measurement is considered cloudy if four or more of the CALIPSO footprints show nonzero extinction. Figure 7 shows the average of cloudy sky water vapor minus clear-sky water vapor as a function of latitude and pressure. Gray shading indicates that the difference is nonzero at the 95% confidence level.

[27] Over most of the upper troposphere and lower stratosphere, cloudy air has higher water vapor than is found in clear skies. The exception is above the tropopause at latitudes equatorward of 20°N. Here, cloudy air has lower water vapor than is found in clear skies. The most likely explanation for this arrangement is that RH controls the effect of clouds on water vapor. In mid and high latitudes, where RH is low, clouds will almost exclusively be associated with convection, and these clouds will tend to moisten [e.g., *Dessler and Sherwood*, 2004]. At low latitudes, where RH is high, in situ formation of clouds can occur, and this will deplete the vapor phase, possibly leading to irreversible dehydration. In addition, it has also been suggested that convection can in some cases also lead to dehydration [e.g., *Sherwood and Dessler*, 2000; *Fueglistaler et al.*, 2009], although this suggestion remains speculative.

4. Conclusions

[28] In this paper, cloud top observations from the CALIPSO instrument are used to study the occurrence of clouds in the Northern Hemisphere summertime lower stratosphere. It is shown that clouds occur frequently in the bottom few kilometers of the stratosphere. At low latitudes, clouds in the stratosphere tend to occur in regions of intense convection, such as the Asian monsoon and over Central America. The 0.1% occurrence contour in these regions reaches 19 km or 420 K potential temperature, about 3 km in altitude or 40 K above the tropopause, while the 2% contour reaches 18 km or 410 K, 2 km or 30 K above the tropopause.

[29] At high latitudes, there is little longitudinal preference for the clouds. From 45°N to 70°N, the 0.1% contour is located around 14 km or 370 K, the 2% contour is located around 13 km or 355 K, and other contours also remain approximately fixed in altitude. The tropopause is descending with increasing latitude, leading to increases in the number of clouds in the stratosphere with latitude.

[30] At low latitudes, stratospheric cloud tops are seen more frequently and at higher altitudes over Asia. In the midlatitudes, however, stratospheric cloud tops are seen more frequently over North America. At high latitudes, the highest frequencies shift back to the Asian sector.

[31] Examining the water vapor fields, it is seen that regions of enhanced cloud occurrence correlate well with regions of enhanced water vapor mixing ratio in the mid and high latitudes. This result is consistent with previous work [e.g., *Dessler and Sherwood*, 2004; *Fu et al.*, 2006; *Gettelman et al.*, 2004] that convection is an important water source for the extratropical lower stratosphere. At low latitudes, the region of enhanced cloud occurrence is not correlated with high water vapor mixing ratio. This is explained by the high relative humidity (RH) found there, which both encourages in situ formation of clouds as well as precludes evaporation of cloud ice.

[32] **Acknowledgments.** This work was supported by NASA grants NNX08AR27G and NNX07AR12G, both to Texas A&M University. CALIPSO data were obtained from the NASA Langley Research Center Atmospheric Science Data Center, and MLS data were obtained from the Goddard Earth Sciences Data and Information Services Center. We thank Mark Vaughn and Dave Winker for their help with the CALIPSO data, and Bill Read and Nathaniel Livesey for their help with the MLS data.

References

Alcala, C. M., and A. E. Dessler (2002), Observations of deep convection in the tropics using the TRMM precipitation radar, *J. Geophys. Res.*, *107*(D24), 4792, doi:10.1029/2002JD002457.

Dessler, A. E., and S. C. Sherwood (2003), A model of HDO in the tropical tropopause layer, *Atmos. Chem. Phys.*, *3*, 2173–2181.

Dessler, A. E., and S. C. Sherwood (2004), The effect of convection on the summertime extratropical lower stratosphere, *J. Geophys. Res.*, *109*, D23301, doi:10.1029/2004JD005209.

Dessler, A. E., S. P. Palm, W. D. Hart, and J. D. Spinhirne (2006a), Tropopause-level thin cirrus coverage revealed by ICESat/Geoscience Laser Altimeter System, *J. Geophys. Res.*, *111*, D08203, doi:10.1029/2005JD006586.

Dessler, A. E., S. P. Palm, and J. D. Spinhirne (2006b), Tropical cloud-top height distributions revealed by the Ice, Cloud, and Land Elevation Satellite (ICESat)/Geoscience Laser Altimeter System (GLAS), *J. Geophys. Res.*, *111*, D12215, doi:10.1029/2005JD006705.

Dessler, A. E., T. F. Hanisco, and S. A. Fueglistaler (2007), Effects of convective ice lofting on H_2O and HDO in the tropical tropopause layer, *J. Geophys. Res.*, *112*, D18309, doi:10.1029/2007JD008609.

Dunkerton, T. J. (1995), Evidence of meridional motion in the summer lower stratosphere adjacent to monsoon regions, *J. Geophys. Res.*, *100*, 16,675–16,688, doi:10.1029/95JD01263.

Fischer, H., et al. (2003), Deep convective injection of boundary layer air into the lowermost stratosphere at midlatitudes, *Atmos. Chem. Phys.*, *3*, 739–745.

Fromm, M. D., and R. Servranckx (2003), Transport of forest fire smoke above the tropopause by supercell convection, *Geophys. Res. Lett.*, *30*(10), 1542, doi:10.1029/2002GL016820.

Fu, R., et al. (2006), Short circuit of water vapor and polluted air to the global stratosphere by convective transport over the Tibetan Plateau, *Proc. Natl. Acad. Sci. U. S. A.*, *103*, 5664–5669, doi:10.1073/pnas.0601584103.

Fueglistaler, S., A. E. Dessler, T. J. Dunkerton, I. Folkins, Q. Fu, and P. W. Mote (2009), The tropical tropopause layer, *Rev. Geophys.*, *47*, RG1004, doi:10.1029/2008RG000267.

Gettelman, A., M. L. Salby, and F. Sassi (2002), Distribution and influence of convection in the tropical tropopause region, *J. Geophys. Res.*, *107*(D10), 4080, doi:10.1029/2001JD001048.

Gettelman, A., D. E. Kinnison, T. J. Dunkerton, and G. P. Brasseur (2004), Impact of monsoon circulations on the upper troposphere and lower stratosphere, *J. Geophys. Res.*, *109*, D22101, doi:10.1029/2004JD004878.
Hanisco, T. F., et al. (2007), Observations of deep convective influence on stratospheric water vapor and its isotopic composition, *Geophys. Res. Lett.*, *34*, L04814, doi:10.1029/2006GL027899.
Hegglin, M. I., et al. (2004), Tracing troposphere-to-stratosphere transport above a mid-latitude deep convective system, *Atmos. Chem. Phys.*, *4*, 741–756.
Hess, P. G. (2005), A comparison of two paradigms: The relative global roles of moist convective versus nonconvective transport, *J. Geophys. Res.*, *110*, D20302, doi:10.1029/2004JD005456.
Hoskins, B. J. (1991), Towards a PV-θ view of the general circulation, *Tellus, Ser. B*, *43*, 27–35.
Jensen, E. J., and L. Pfister (2004), Transport and freeze-drying in the tropical tropopause layer, *J. Geophys. Res.*, *109*, D02207, doi:10.1029/2003JD004022.
Jost, H.-J., et al. (2004), In-situ observations of mid-latitude forest fire plumes deep in the stratosphere, *Geophys. Res. Lett.*, *31*, L11101, doi:10.1029/2003GL019253.
Livesey, N. J., M. D. Fromm, J. W. Waters, G. L. Manney, M. L. Santee, and W. G. Read (2004), Enhancements in lower stratospheric CH_3CN observed by the Upper Atmosphere Research Satellite Microwave Limb Sounder following boreal forest fires, *J. Geophys. Res.*, *109*, D06308, doi:10.1029/2003JD004055.
Livesey, N. J., et al. (2007), Version 2.2 Level 2 data quality and description document, *JPL D-33509*, 109 pp., Jet Propul. Lab., Pasadena, Calif.
Newman, P. A., and J. E. Rosenfield (1997), Stratospheric thermal damping times, *Geophys. Res. Lett.*, *24*, 433–436, doi:10.1029/96GL03720.
Poulida, O., R. R. Dickerson, and A. Heymsfield (1996), Stratosphere-troposphere exchange in a midlatitude mesoscale convective complex: 1. Observations, *J. Geophys. Res.*, *101*, 6823–6836, doi:10.1029/95JD03523.
Sherwood, S. C., and A. E. Dessler (2000), On the control of stratospheric humidity, *Geophys. Res. Lett.*, *27*, 2513–2516, doi:10.1029/2000GL011438.
Suarez, M. J., et al. (2008), The GEOS-5 Data Assimilation System—Documentation of versions 5.0.1, 5.1.0, and 5.2.0, *NASA Tech. Rep., TM-2008-104606*, vol. 27, 97 pp.
Wang, P.-H., P. Minnis, M. P. McCormick, G. S. Kent, and K. M. Skeens (1996), A 6-year climatology of cloud occurrence frequency from Stratospheric Aerosol and Gas Experiment II observations (1985–1990), *J. Geophys. Res.*, *101*, 29,407–29,429, doi:10.1029/96JD01780.
Wang, P. K. (2003), Moisture plumes above thunderstorm anvils and their contributions to cross-tropopause transport of water vapor in midlatitudes, *J. Geophys. Res.*, *108*(D6), 4194, doi:10.1029/2002JD002581.
Winker, D. M., M. A. Vaughan, A. Omar, Y. Hu, K. A. Powell, Z. Liu, W. H. Hunt, and S. A. Young (2009), Overview of the CALIPSO mission and CALIOP data processing algorithms, *J. Atmos. Oceanic Technol.*, doi:10.1175/2009JTECHA1281.1, in press.
Zipser, E. J., D. J. Cecil, C. Liu, S. W. Nesbitt, and D. P. Yorty (2006), Where are the most intense thunderstorms on Earth?, *Bull. Am. Meteorol. Soc.*, *87*, 1057–1071, doi:10.1175/BAMS-87-8-1057.

A. E. Dessler, Department of Atmospheric Sciences, Texas A&M University, TAMU 3150, College Station, TX 77843-1427, USA. (adessler@tamu.edu)